Low-*P* metamorphism following a Barrovian-type evolution. Complex tectonic controls for a common transition, as deduced in the Mondoñedo thrust sheet (NW Iberian Massif)

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Abstract

The Mondoñedo thrust sheet has been studied to investigate the complex dynamic relationships that may be involved in the development of low- and medium-*P* metamorphic domains. This unit underwent an initial medium-*P* event during the initial stages of Variscan convergence, related to crustal thickening. Subsequently, the thrust sheet evolved to a low-*P* baric type of metamorphism, related to syn-convergence thinning and exhumation. Its footwall, cropping out in two tectonic windows, registered a different evolution, with a low-*P* history that evolved from low- to high-*T* under a high geothermal gradient. Several different $P-T$ paths of the Mondoñedo thrust sheet and its relative autochthon are traced and interpreted according to the structural evolution of the area. Following the initial crustal thickening, two main syn-convergence extensional shear zones developed. One of them occurs in the hangingwall, whereas the other affects the footwall unit. Both extensional shear zones were contemporaneous with ductile thrusting in the inner parts of the thrust sheet, and their activity is viewed as a consequence of the need for gravitational re-equilibration within the orogenic wedge.

The most commonly accepted models of tectonothermal evolution in regions of thickened continental crust assume that low-*P* metamorphism is essentially a late phenomenon, and is linked to late-orogenic tectonic activity. In the Mondoñedo thrust sheet, our conclusions indicate that low-*P* metamorphism may also develop during convergence, and that this may occur in at least two cases. One is tectonic denudation of an allochthonous unit during its emplacement, and the other, thinning and extension at the footwall unit of an advancing thrust sheet. As a consequence, the low-*P* evolution may show different characteristics in different units of an orogenic nappe pile.

Keywords: Variscan metamorphism; Barrovian to low-*P* evolution; NW Iberian Massif

1. Introduction

Thermal models of orogenic regions characterized by crustal thickening have shown that their thermal evolution follows rather systematic patterns. For intermediate values of the more common parameters intervening in the thermal history, mesocrustal domains

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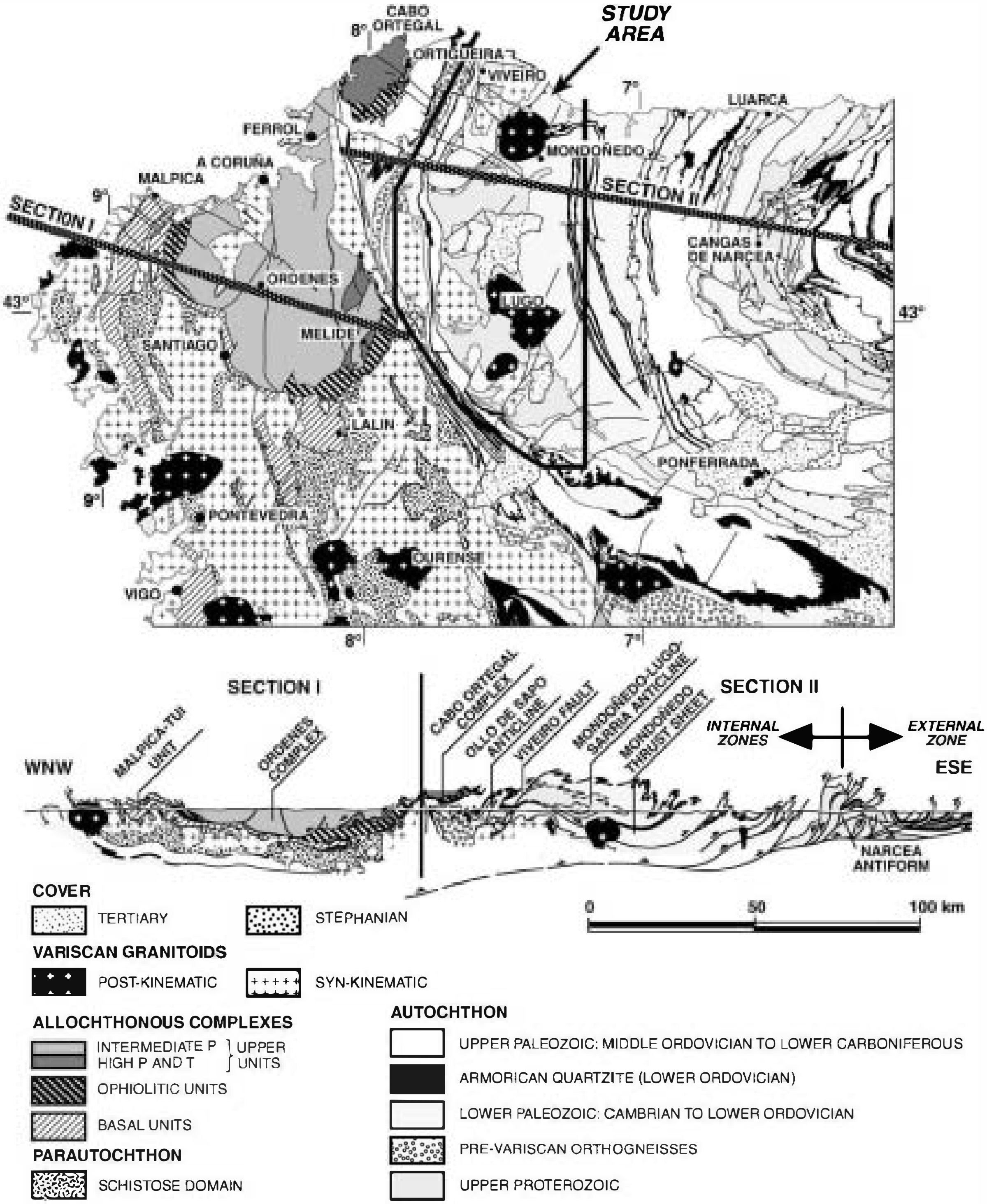

Fig. 1. Geological map and cross section of the NW Iberian Massif, with location of the study area.

undergo a medium-pressure metamorphic evolution (England and Thompson, 1984). In fact, this Barrovian-type evolution has been considered an indicative for collision tectonics (Thompson and England, 1984). However, for most middle and lower crustal levels, the medium-P metamorphic history follows from a previous one of high-P roughly synchronous with the maximum burial. Nevertheless, because of the inferred clockwise shape of the $P-T-t$ paths developed in the medium to higher P parts of the belt, evidence for the early maximum P segment of the $P-T-t$ path is usually obliterated by the subsequent recrystallization at peak T, specially when the $P-T$ path reach a high-T thermal peak.

Numerical experiments show that most of the $P-T$ paths inside a thickened crust tend to approach the low-P metamorphic field after the thermal peak and during strong decompression. When exhumation is only due to erosion, England and Thompson (1984) have suggested that the low-P field is not reached by the common $P-T$ paths. But, as a late event of low-P is the norm rather than the exception in collisional orogenic belts, one has to conclude that erosion is not the only process responsible for the low-P metamorphic gradients. Abundant magmatism may help the trajectories to enter the low-P domain of the $P-T$ space, but late-orogenic extension has also been claimed to explain broad regions with low-P metamorphism (England, 1987; Thompson and Ridley, 1987). Many regional contributions have described the quick exhumation of deep crustal sectors in the footwall of large extensional detachments (Sandiford, 1989; Platt, 1993; Ruppel, 1995). These sectors appear nowadays as dome-shaped, low-P plutono-metamorphic complexes (Dewey, 1988; Reinhardt and Kleeman, 1994; Escuder Viruete et al., 1994, 2000). Both plutono-metamorphism and extension are the mechanisms most commonly used to explain the low-P metamorphism in collisional belts.

This contribution presents a study carried out in the Mondoñedo thrust sheet and its authochthon, in the NW of the Iberian Massif (Fig. 1), which forms part of the Variscan orogenic belt. Our results indicate that regional low-P metamorphism develop in sectors of collisional orogenic wedges that register a rich and complex structural evolution. The low-P event occurred there both in the hangingwall and the footwall of a huge contractional structure. In the thrust sheet itself, the low-P metamorphism took place after a medium-P event clearly related to crustal thickening, but in the footwall, the low-P metamorphism is the only significant event of recrystallization. Moreover, the complex structural and metamorphic relationships established between the upper and lower blocks of the Mondoñedo thrust fault give insights into the evolution of the Variscan orogenic wedge with time. Histories like the one described here may be relatively common in orogenic belts, though could they have been remained unnoticed in many cases.

2. Geological setting

A first deformation episode of recumbent folding with E vergence characterizes the Variscan orogeny in the internal parts of the NW Iberian Massif. This was followed by ductile and brittle thrusting toward the E, and subsequent open steep folding (Fig. 1).

The Mondoñedo nappe is a large crystalline thrust sheet, formed essentially by low to high-grade metasediments ranging in age from Upper Proterozoic to Lower Devonian. Metapelites are the most common lithologies, followed by quartzites and, less abundant, carbonates. The thickness of the sedimentary succession varies, reaching at most 6000 m, but due to recumbent folding, its present structural thickness roughly doubles the sedimentary one. To the W, the thrust sheet is bounded by the Viveiro fault, a west-dipping normal fault cutting across the nappe and its relative autochthon (Figs. 2 and 3). Late folding allowed the present-day preservation from erosion of 10 to 12 km of the thrust sheet in the core of an open synform.

The structure of the hangingwall unit and the basal ductile shear zone has been described by Bastida and Pulgar (1978), Martínez Catalán (1985), Bastida et al. (1986) and Aller and Bastida (1993). The significance of the Mondoñedo nappe in the structural evolution of the Variscan belt of NW Spain is discussed in Pérez-Estaún et al. (1991).

What makes the nappe a very interesting structure is the fact that it can be studied from its front, that runs along 200 km (Marcos, 1973; Pérez-Estaún, 1978), to rather internal areas some 65 km to the west, where its relative autochthon is exposed in two tectonic windows, named Xistral and Monte Carballosa (Fig. 2).

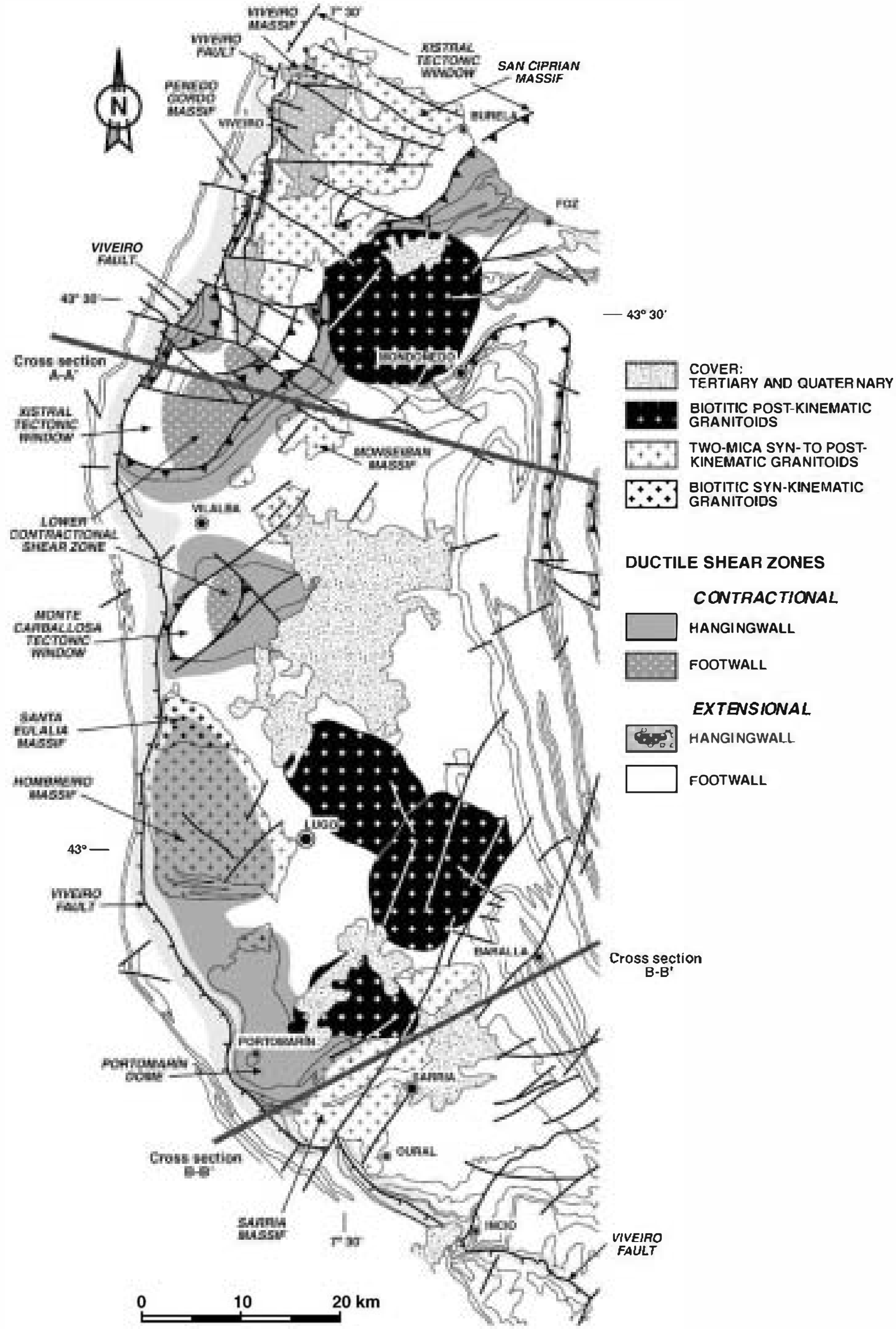

Fig. 2. Map showing the distribution of ductile shear zones in the hangingwall and footwall to the Mondoñedo thrust sheet.

There, the succession is thought to range form Upper Proterozoic to Lower Cambrian, but the quartzites dominate over the metapelites and metagreywackes. A careful review of the hangingwall unit and new mapping of the footwall unit in the tectonic windows has shown a complex picture of ductile shear zones with different kinematics and partially overlapping in time. The structural evolution will be presented in a companion paper (Martínez Catalán et al., in press), whereas this contribution deals with the metamorphic evolution.

The first study of metamorphism in the nappe and surrounding areas was carried out by Capdevila (1969). It included a map of the metamorphic zones and a systematic description of the mineral assemblages in the metapelites. Though it was later put in an updated regional context by Bastida et al. (1986), Capdevila's contribution remained the only work to approach the metamorphic evolution at the scale of the whole Mondoñedo nappe. Its main result is that the regional metamorphism was of the low-*P* intermediate type of Miyashiro (1961). Nevertheless, the author describes the local appearance of kyanite, with or without coexisting andalusite that he interpreted as related to individual fault zones, such as the Viveiro fault. The kyanite-bearing metapelites apparently associated to this fault have received the attention of Martínez et al. (1996) and Reche et al. (1998a,b). The latter carried out quantitative thermobarometric estimations in restricted areas of the nappe.

However, the tectonothermal evolution of the nappe and its relative autochthon had not been approached following a modern perspective that would pay attention to the structural evolution of both the hangingwall and footwall units, and that try to interpret the P–T paths in terms of the evolving macrostructures. This is the aim of this contribution, and we will show that the metamorphism depicts a complex evolution that, even reaching high temperatures in both units, followed very different paths. Whereas a typical medium-*P* metamorphism of Barrovian affinities characterizes the hangingwall unit, a low-*P* evolution took place below the Mondoñedo thrust sheet. The model envisaged to explain these differences may be applied to other sectors and orogenic belts where an alternation of medium- and low-pressure metamorphic regimes occur.

3. Synmetamorphic shear zones

The first structures developed in the study area were large recumbent folds. These folds reflect an episode of crustal shortening and thickening, with the deformation widely distributed and developing a regional cleavage affecting all the metasediments. The associated regional metamorphism was of intermediate pressure (Miyashiro, 1961) with kyanite–sillimanite, so common in mesocrustal levels of many orogenic belts (Thompson and England, 1984). The metamorphic zonation, of Barrovian type, includes chlorite, biotite, garnet, staurolite–kyanite, sillimanite and sillimanite–orthoclase (Fig. 4). The isograds crosscut the recumbent folds (Capdevila, 1969), but they were deformed by the subsequent ductile shear zones, and overprinted by the thrust and normal faults (Fig. 3).

The most important recumbent fold is the Mondoñedo–Lugo–Sarria anticline (MLS). Its core, occupied by Upper Proterozoic metapelites and metagreywackes (Fig. 1), has a mean thickness of 6 km along most of the fold, as deduced from cross sections constructed by down-plunge projection of the hinges and limbs of the second-order folds. To the W and SW, however, the normal and reverse limbs progressively approach each other, and the fold core becomes less than 1 km thick (Fig. 3). The thinning of the fold nappe toward its internal parts is reflected also in the thickness of the Paleozoic formations on both limbs and in the width of the metamorphic zones (Figs. 3 and 4), and is due to the superposition of two ductile shear zones with opposite senses of movement.

One is the basal shear zone of the Mondoñedo nappe, a 3–3.5-km-wide zone of ductile deformation with top to the E motion, that occurs at the base of the thrust sheet (Bastida and Pulgar, 1978; Bastida et al., 1986; Aller and Bastida, 1993). This is interpreted as a contractional structure equivalent to a ductile thrust, and it is assumed that the allochthonous sheet moved without much internal deformation, except at the basal zone and, once it had reached relatively high crustal levels, motion occurred over a brittle thrust fault.

The other is a low-dipping shear zone with top to the W motion, developed in the upper parts of the MLS anticline (Figs. 2 and 3). This zone of deformation is clearly superimposed to the normal limb of the MLS anticline. It shows an associated subhorizontal crenulation cleavage or a new schistosity, as well as a

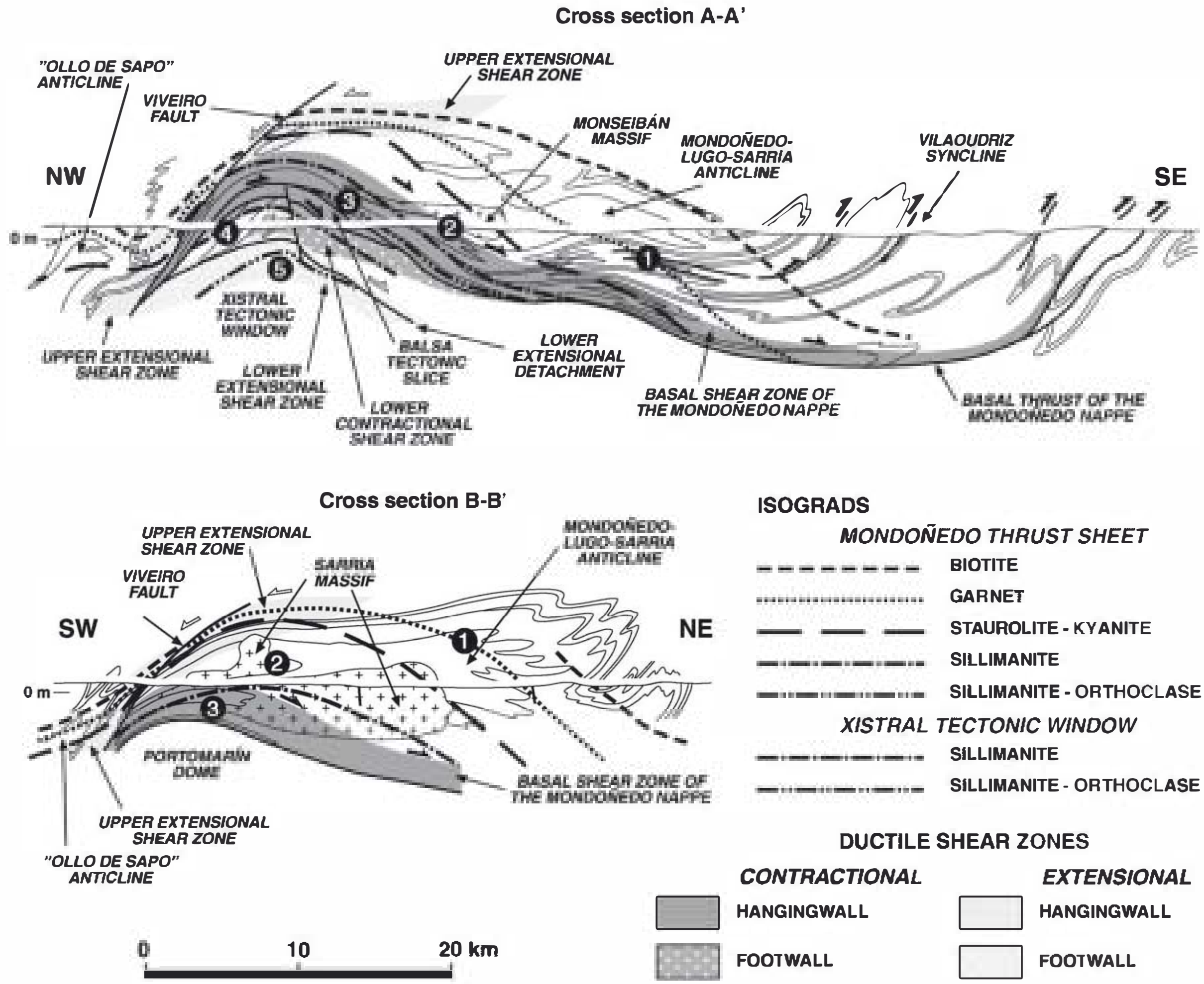

Fig. 3. Two general sections across the Mondoñedo nappe, showing the shear zones and the metamorphic isograds. Arrows show movement of contractional (black) and extensional (white) shear zones and faults. For location, see Fig. 2. Numbers 1 to 5 refer to the location of the $P-T$ paths shown in Fig. 9.

mineral lineation striking E–W. Its maximum thickness is estimated in 2 km, and decreases progressively to the S of Oural (Fig. 2). This shear zone has a subtractive character, as deduced from the thinning of the normal limb and core of the MLS anticline, and also of the metamorphic zones (Fig. 3, section B-B′), and will be called the upper extensional shear zone.

The two shear zones mentioned occur in the hangingwall to the Mondoñedo thrust fault, but synmetamorphic shear zones have been identified also at its footwall. Relics of the basal shear zone of the Mondoñedo nappe have been preserved in the upper part of the footwall unit (Figs. 2 and 3). In addition, a zone of high strain with kinematic criteria indicating top to the E shearing has been identified in the two tectonic windows. Geometrically, this lower contractional shear zone merges with the Mondoñedo thrust upward to the E, implying that prior to late folding, it was more inclined to the W than the Mondoñedo thrust itself. Because its dip was toward the upthrown block, it is interpreted as a ductile thrust.

However, the most important ductile shear zone occurs in the deep parts of the Xistral tectonic window. A detachment fault underlies a thick quartzitic formation and, below, intense shearing affects the underlying schists, paragneisses and migmatites outcropping at the core of the open antiform that is at the origin of the tectonic window (Figs. 2 and 3, section A-A′). Shearing shows a top to the E sense of motion but, in this case, structural criteria suggest that its flow plane had an original dip to the E, i.e., to the downthrown block. Furthermore, a subtractive metamorphic jump that will be described below supports that this is an extensional structure, that will be referred to as the lower extensional shear zone. Its upper limit is the lower extensional detachment (Fig. 3, section A-A′).

All of the structures described above were active during different stages of nappe emplacement. Later on, the Viveiro fault overprinted both the hangingwall and footwall units of the Mondoñedo thrust sheet (Figs. 2 and 3). This is a normal brittle fault dipping 40–60°W that also has an associated narrow ductile shear zone (not shown in the maps and cross sections). Minor recumbent folds with W vergence and weakly curved hinges, and S–C or ECC microstructures (Platt, 1984), indicate a top to the W motion on the fault, with a slight right-lateral component.

4. Metamorphic zones and isograds

The regional distribution of metamorphic zones is shown in Figs. 4 and 5. They have been traced through the study of mineral parageneses in significant geological sections. A fundamental conclusion is the difference in the metamorphic zonation between the thrust sheet and its relative autochthon. In the former, the evolution is one of medium-*P*, essentially Barrovian, but with a continuation towards temperatures somewhat higher than those registered in the type area of Scotland (Yardley, 1989). This permits to map the classical Barrovian zones, widely described in the literature. Conversely, in the footwall unit, the evolution is of lower pressure, with a less evident metamorphic zonation that requires a specific selection.

In the hangingwall unit, the mapped zones are, from top to bottom, those of chlorite, biotite, garnet, staurolite–kyanite, sillimanite and sillimanite–orthoclase. The upper three zones occur in the central and eastern sectors of the Mondoñedo nappe, and the isograds separating them show a roughly N–S attitude (Fig. 4). The staurolite–kyanite zone occupies most of the western part of the nappe (Figs. 4 and 5). Commonly, in Barrovian terranes, the staurolite and kyanite zones can be separated, the latter being the deeper (Yardley, 1989). However, in the Mondoñedo thrust sheet, coexistence of both phases occurs immediately below the garnet zone. That is, the first staurolite-bearing schists also contain kyanite (together with chloritoid, chlorite, white mica, quartz, plagioclase and ilmenite). This prevents the separation of the two zones. Probably, the quick appearance of kyanite indicates a metamorphic evolution in the higher-pressure domain of the medium-*P* metamorphism. To the W, the metamorphic zonation of the thrust sheet is truncated by the Viveiro fault.

The first sillimanite zone, characterized by schists and paragneisses with sillimanite and muscovite, crops out in three areas: as a narrow strip to the E of the Xistral tectonic window, to the south of the Hombreiro granitic massif and in the Portomarín dome (Figs. 4 and 5). The second sillimanite zone, characterized by migmatitic paragneisses with sillimanite and orthoclase without primary muscovite, is restricted to an even narrower strip to the E of the Xistral tectonic window. These paragneisses represent the highest metamorphic grade identified in the area; no outcropping rocks ever reached the sillimanite–cordierite–orthoclase or higher temperature zones.

The metamorphic succession of the footwall unit developed under lower baric conditions. In similar domains, a generally accepted model for the metamorphic zonation is lacking, and different zones have been described in different regions (Miyashiro, 1961; Harte and Hudson, 1979; Yardley, 1989). Consequently, we must first consider what zones can be mapped. Beginning from the deepest parts, in the core of the Xistral tectonic window, paragneisses without primary muscovite and clear evidence of partial melting can be ascribed to the sillimanite–orthoclase zone (Fig. 5). Above, the paragneisses and schists are characterized by sillimanite and syntectonic primary muscovite, without associated partial melting, so that they must represent the first sillimanite zone.

More problematic is the zonation of the overlying levels, corresponding to the intermediate temperatures. For similar metamorphic conditions, it is common the separation of the cordierite and andalusite zones (Harte and Hudson, 1979), the first being of lower temperature. However, in our case, the medium-*T* metasedi-

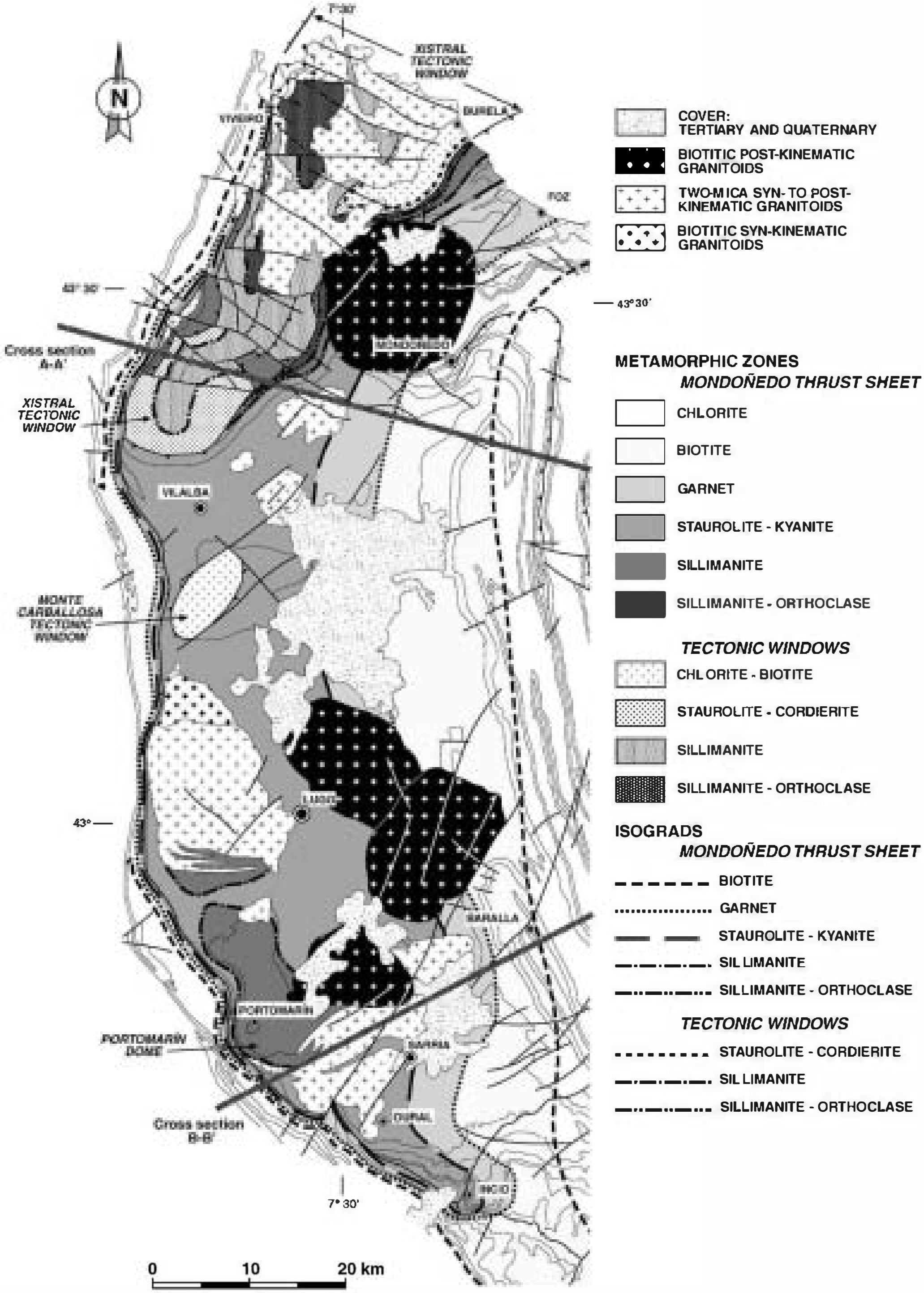

Fig. 4. Map of metamorphic zones and isograds in the Mondoñedo thrust sheet and its footwall unit.

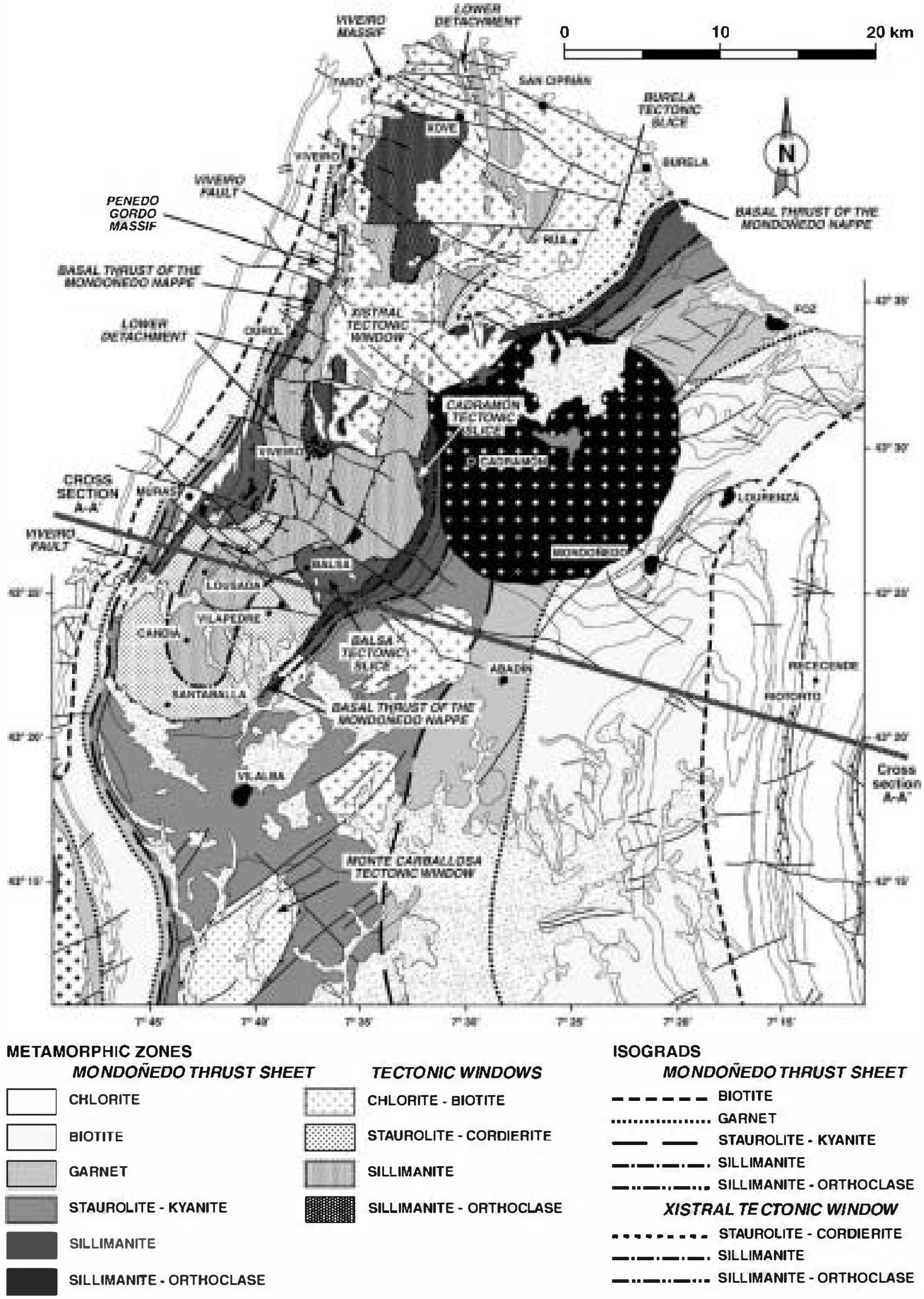

Fig. 5. Detailed map of metamorphic zones and isograds in the tectonic windows and surrounding parts of the Mondoñedo thrust sheet.

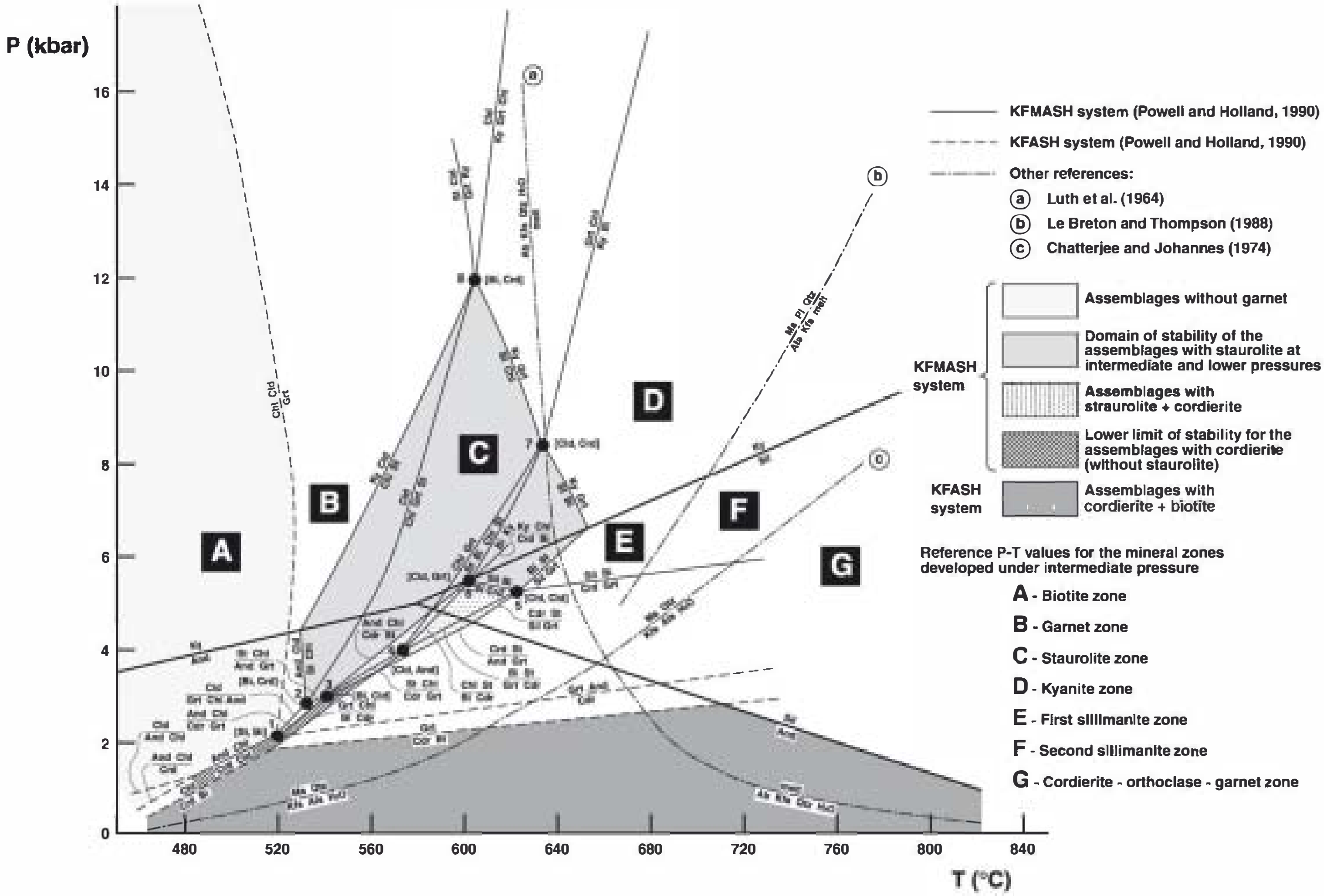

Fig. 6. Petrogenetic grid for metapelites used to constrain the metamorphic evolution of the Mondoñedo thrust sheet and its footwall unit. The grid includes a collection of significative reactions in the KFMASH and KFASH systems (Powell and Holland, 1990), as well as other important equilibria in metapelites (Luth et al., 1964; Chatterjee and Johannes, 1974; Le Breton and Thompson, 1988). Some significant stability domains and reference $P-T$ conditions for the characteristic mineral zones of the intermediate-P metamorphism are also shown.

ments of the footwall unit may contain staurolite in addition to cordierite, and the andalusite is not common at all. For this reason, a single staurolite–cordierite zone has been defined

by the petrogenetic grid of Powell and Holland (1990) for metapelites that shows the possible coexistence of staurolite and cordierite at low pressures. This petrogenetic grid will be used to discuss the P–T paths of both the hangingwall and footwall units (Fig. 6).

Low-T metasediments characterize the whole Monte Carballosa tectonic window and also occur to the S and SW of Burela, in a tectonic slice in the NE part of the Xistral tectonic window (Fig. 5). The rocks there are mostly quartzites, which evidently does not help to establish a detailed metamorphic zonation. For that reason, and considering their mineral assemblage, they have been ascribed to a chlorite–biotite zone.

5. *P*–*T* paths

One of the main tools to analyse the metamorphic evolution of a complex area are the P–T paths followed by the most significant

a marked dependence of the dynamic history of each orogenic sector (England and Thompson, 1984; Thompson and England, 1984; Davy and Gillet, 1986; England, 1987; Thompson and Ridley, 1987). After a careful petrographic analysis, two techniques can be used to establish the P–T paths. One is the use of geothermometers and geobarometers applied to mineral assemblages in equilibrium, formed along successive stages. The other uses petrogenetic grids to analyse the reactional space where the actual parageneses are stable.

By using thermobarometry, individual paths can be traced with precision (Arenas et al., 1995, 1997; Escuder Viruete et al., 1997, 2000), but this is a very slow task when one has to interpret a wide region and just wants to catch the essential features of its tectonothermal evolution. The alternative approach will be applied here, using a petrogenetic grid which, in any case, permits to compare rather accurately the temperatures, though is somewhat less precise for the pressures, mainly inside the stability domain of kyanite or the high-T part of sillimanite.

We have used the petrogenetic grid of Powell and Holland (1990) for the pelitic system (KFMASH), adding several reactions of the system KFASH obtained by the same authors, and the curves of Luth et al. (1964), Le Breton and Thompson (1988) and Chatterjee and Johannes (1974), for the minimum melting in the hydrous granitic system, and for the disappearance of muscovite (Fig. 6). Also, several domains of mineral stability meaningful for the Mondoñedo

P–T reference conditions for mineral zones typical of medium-P metamorphism. Independently of the accuracy in the establishment of the P and T conditions, the grid of Powell and Holland (1990) may be used as a tool for comparison, because of its internal coherency, obtained by rigorous thermodynamic calculations in the system of reference. In order to establish comparisons, this grid is a better tool than a group of reactions deduced from experiments carried out with different initial compositions.

In the rest of this chapter, the P–T paths of the most significant

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tion is based on a detailed petrographic study of these lithologies, together with the interpretation of the tectonic fabrics (foliations and lineations) and their relationships with the macrostructures.

5.1. P–T paths in the hangingwall unit

The upper parts of the Mondoñedo

characterized by the zones of chlorite, biotite and garnet. Considering the mineral assemblages of the two latter zones, and assuming a geothermal gradient of 20 °C/km during burial, two approximate P–T paths have been traced taking into account the reaction Chl + Cld = Grt (KFASH system, Fig. 7, paths A and B).

More detail can be obtained for the metamorphic evolution of the schists on the normal limb of the MLS anticline to the W of Vilalba and farther to the S, around Incio (Fig. 4). These schists occupy the upper part of the staurolite–kyanite zone, and may be divided in two main types, according to the presence or absence of kyanite. The latter have a higher Fe-content, and their paragenesis includes muscovite, biotite or chloritoid, chlorite, quartz, plagioclase, garnet, staurolite and ilmenite. Temporary relationships among chloritoid, garnet and staurolite show a typical Barrovian reactional history.

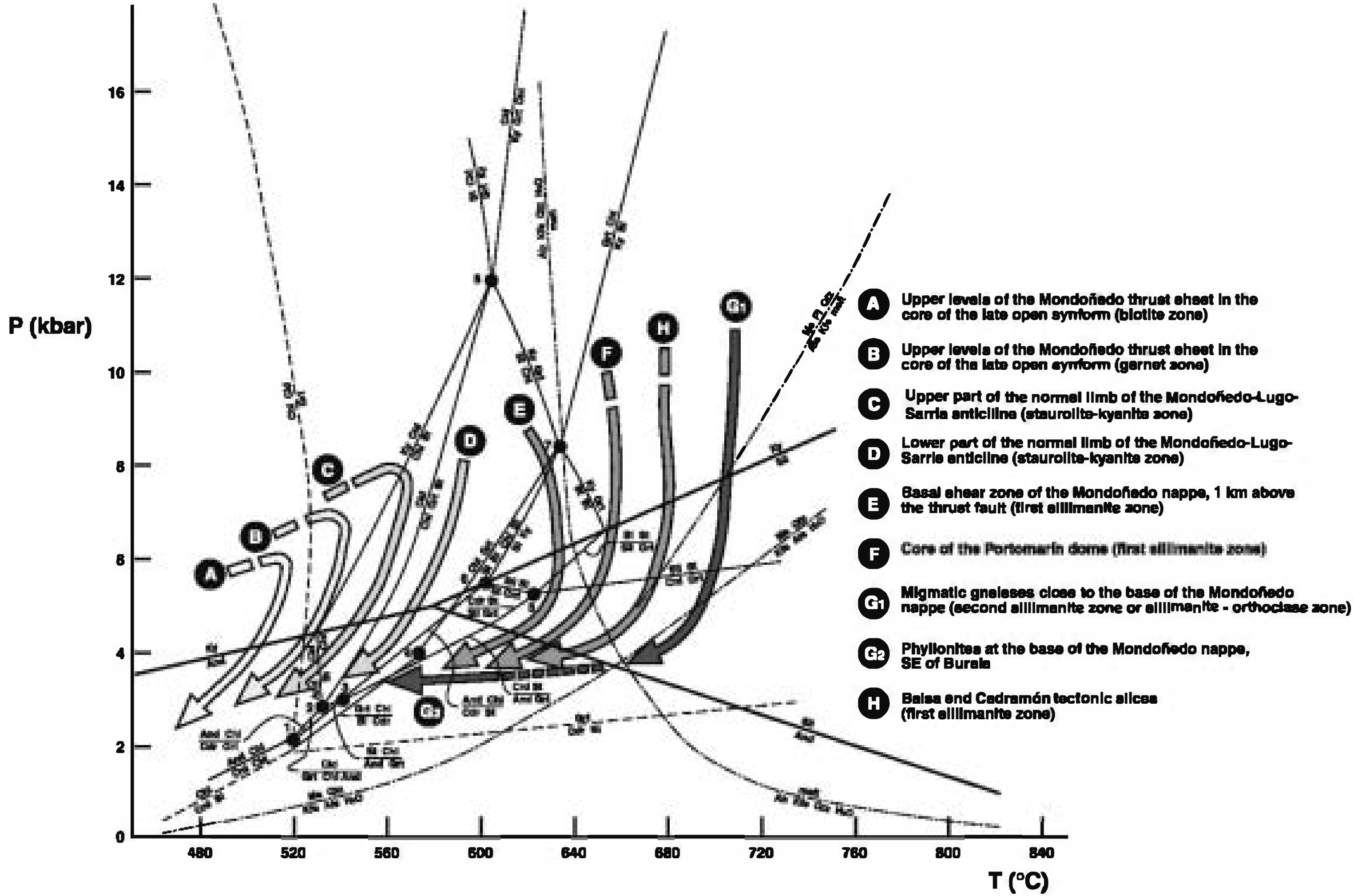

Fig. 7. $P-T$ paths deduced for the different structural levels of the Mondoñedo thrust sheet.

The kyanite-bearing schists developed from Al-rich compositions, also relatively rich in Mg. Their paragenesis includes muscovite, chlorite, quartz, plagioclase, kyanite, staurolite, chloritoid and ilmenite, but biotite and garnet are absent. This assemblage suggests that the univariant reaction Ky + Cld = Chl + St (KFMASH system) resulted in a stable association in equilibrium with chlorite + chloritoid + kyanite + staurolite. Considering that the slightly late character of the staurolite could imply that the equilibrium of the reaction was finally exceeded,
traced to the right of the reaction curve (Fig. 7, path C). The pressure is rather imprecise in this path, due to the lack of reactions giving baric information in the kyanite field for
path, we have used the same thermal gradient as for the garnet and biotite zones, but clearly, the pressure peak must be considered as a lower limit. In any case, the systematic absence of garnet in the assemblages with kyanite would place the baric peak below the invariant point 8 (Figs. 6 and 7). This path underwent exhumation through temperatures lower than those of its baric peak, given the lack of significant
associated to decompression.

For intermediate sectors of the staurolite–kyanite zone, the common association contains biotite, garnet, staurolite, and occasionally chloritoid only as micro-inclusions in garnet. Furthermore, andalusite is common as late-kinematic porphyroblasts. These are typical schists of the staurolite (or staurolite–kyanite) zone, that previously prograded through chloritoid and garnet and subsequently were decompressed and chilled syn-kinematically through the andalusite field. Consequently, the P–T path (D in Fig. 7) has been traced in the medium P and T domain of the staurolite stability, at higher T than the stability limit of chloritoid (equilibrium Cld = Chl + Grt + St) and entering the andalusite field
Probably, the retrograde path was always above the invariant point 2, considering the absence of cordierite. On the other hand, the rocks of the staurolite–kyanite zone never exceeded significantly
the triple point of the Al-silicate polymorphs, because of the absence of sillimanite during decompression. In that way, they probably never exceeded the equilibrium Chl + Grt = Bi + St.

Deeper in the nappe structure, the schists of the first sillimanite zone to the E of the Xistral tectonic window (Fig. 5) contain a syn-kinematic assemblage with garnet, staurolite, biotite and sillimanite, followed by andalusite porphyroblasts (up to 15 cm long, and parallel to the stretching lineation). The P–T path (E in Fig. 7) can be traced very precisely showing a strong decompression and intersecting the equilibria Chl + Grt = Bi + St and Bi + St = Sil + Grt, to finally
the andalusite stability field.
traced for the schists and paragneisses of the first sillimanite zone in the Portomarín dome (Fig. 4). These rocks may contain syn-kinematic biotite, garnet, staurolite, kyanite and sillimanite, and late-kinematic andalusite, so that it is quite possible that their P–T path followed higher temperatures than the equilibria Grt + Chl = Ky + Bi, St + Bi = Ky + Grt, and Bi + St = Sil + Grt (Fig. 7, path F).

The second sillimanite zone was reached at the base of the thrust sheet to the E of the Xistral tectonic window (Fig. 5). The rocks are migmatitic paragneisses that may include quartz, plagioclase, biotite, garnet, staurolite, orthoclase, sillimanite and andalusite. However, the most recrystallized, biotite-poor anhydrous rocks contain quartz, plagioclase, sillimanite and orthoclase in the assemblage representing the thermal peak. The P–T path shows a strong decompression through the sillimanite stability field,
the curves of disappearance of muscovite and finally loosing temperature through the andalusite domain (Fig. 7, path G1). This path may be continued to lower temperatures studying the phyllonites adjacent to the basal thrust fault at the coastal section, SE of Burela (Fig. 5). The phyllonites resulted from exhumation and chilling of the migmatitic paragneisses during thrusting. Thermobaric estimations carried out in these fault rocks (Dallmeyer et al., 1997) suggest conditions of less than 5 kbar (maximum content of Si in white mica = $Si_{3.18}$) and temperature around 500 ± 30 °C (plagioclase–muscovite equilibrium) for the end of the process (Fig. 7, path G2).

To the E of the Xistral tectonic window, a couple of imbricates occurs between the Mondoñedo
sheet, and its footwall unit, the Balsa and Cadramón tectonic slices (Figs. 5 and 3, section A-A′). Both their stratigraphy and metamorphic evolution indicate that they are fragments detached from the thrust sheet during its emplacement. They clearly belong to the first sillimanite zone, but their P–T conditions were somewhat intermediate between the basal migmatites

and the paragneisses of the Portomarín dome (Fig. 7, path H).

5.2. *P–T paths in the footwall unit*

In the footwall to the Mondoñedo lowest *P–T* conditions are those of the chlorite–biotite zone, preserved in the Burela tectonic slice and the Monte Carballosa tectonic window (Fig. 5). This and the underlying staurolite–cordierite zone are hard to trace with precision because they occur in an area dominated by quartzites. In Burela, only chlorite and white mica occur in the rare semipelites, but in Monte Carballosa, post-kinematic biotite porphyroblasts suggests a somewhat higher grade. The *P–T* path representing the evolution of both areas is shown in Fig. 8 (path I).

Around Vilapedre, to the S of the Xistral tectonic window (Fig. 5), the schists of the staurolite–cordierite zone contain quartz, plagioclase, white mica, chlorite, biotite and cordierite. Cordierite appears as late- to post-kinematic porphyroblasts and their occurrence indicates a pronounced heating during final stages of the development of the regional foliation (Fig. 8, path J).

In the upper parts of the footwall unit, the general absence of garnet (just one sample with possible weathered garnets) makes the drawing of paths I and

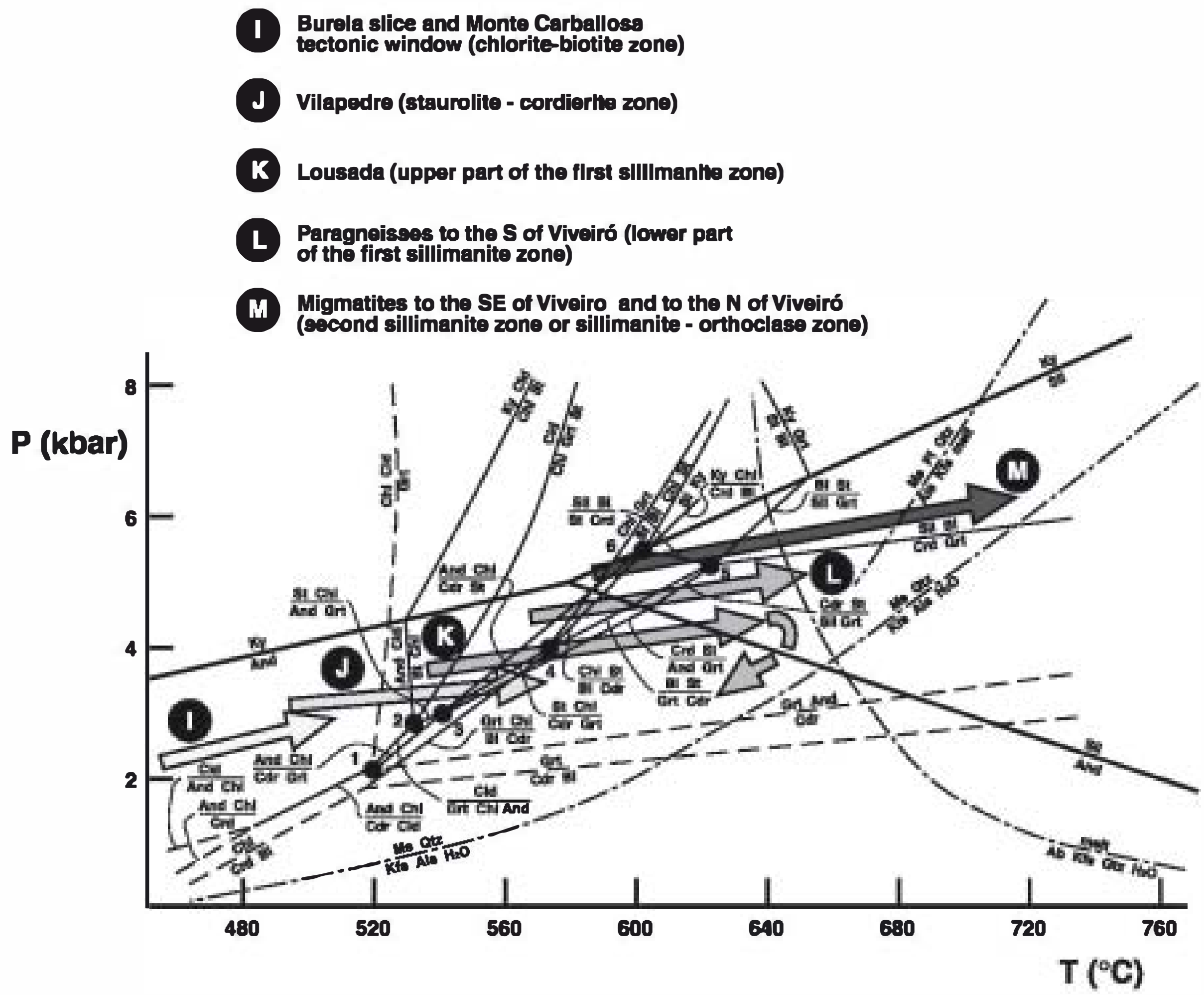

Fig. 8. *P–T* paths deduced for the different structural levels of the footwall unit.

J uncertain. Strictly, both arrows should be traced below the invariant point 1, where the mineral assemblage includes garnet. However, the pressures would then be so low that are incompatible with the structural data. A thickness of at least 10 km is preserved today in the central parts of the Mondoñedo
and some erosion must be expected since the Carboniferous. This is incompatible with the approximately 8 km corresponding to the 2 kbar of invariant point 1. The absence of garnet may have a compositional explanation but anyway, the maximum pressure sustained by these rocks should have been low.

The schists of the upper part of the first sillimanite zone around Lousada (Fig. 5) may contain biotite, staurolite, cordierite, sillimanite, some garnet and primary muscovite. The cordierite grew later than garnet and staurolite. Still later growth of sillimanite represents the highest temperature reached in the area. The arrow has been traced crossing the stability fields
assemblages with staurolite and staurolite + cordierite, reaching finally
the curve of partial melting (Fig. 8, path K). It is possible that these schists developed an assemblage with garnet + staurolite + biotite + chlorite + cordierite during their evolution. Then, the $P-T$ path would pass through the invariant point 4. Some cordierite porphyroblasts are syn-kinematic, but the largest are post-kinematic, so that their growth may be linked to the final
with a slight decompression, following a $P-T$ path that did not reach the curve of disappearance of muscovite.

The lower part of the first sillimanite zone can be studied to the S of Viveiró, in the centre of the Xistral tectonic window (not to be confused with Viveiro, to the N and in the coast, that gives its name to the Viveiro fault; see Fig. 5). The rocks are paragneisses with sillimanite, biotite and primary muscovite, but their strong recrystallization has overprinted any trace of early phases of their thermal history. Garnet has never been found in these rocks, which hinders the precise tracing of their $P-T$ path. In any case, the evolution is one of low pressure, and has been drawn parallel to that of the schists around Lousada but entering somewhat more inside the stability field
sillimanite (Fig. 8, path L). The absence of partial melting and the systematic occurrence of primary muscovite limit the progradation of this path into the sillimanite field.

The migmatites occurring to the E of Viveiro and also to the N of Viveiró (Fig. 5) are the deepest lithologies outcropping in the footwall unit. Their mineral assemblage was developed during the thermal peak, and no previous mineral phases have been preserved. They contain sillimanite, biotite, orthoclase and, occasionally, garnet. Partial melting is widespread and all the white mica is secondary. The final progradation of the $P-T$ path has been traced inside the sillimanite field,
disappearance of muscovite (Fig. 8, path M). For earlier stages, the path cannot be established due to the practical absence of pre-peak phases. It is doubtful if the arrow went below the triple point of Al-silicate polymorphs or along the stability field
existence of the lower extensional shear zone and an associated detachment (Fig. 3, section A-A′) suggests a structural thickness above the migmatites more important before the beginning of extension. This supports the tracing of the $P-T$ path M above that of path L (Fig. 8) and, consequently, around the triple point. The absence of kyanite and the scarcity of garnet suggests that, in any case, the pressures were not too high.

6. Interpretation of the $P-T$ paths

A synthesis of the $P-T$ evolution of the study area is shown in Fig. 9. Paths 1, 2 and 3 correspond to the upper, middle and lower parts of the Mondoñedo
sheet and paths 4 and 5 to its relative autochthon, above and below the lower extensional detachment respectively. Fig. 3 depicts the structural location of the $P-T$ paths shown in Fig. 9.

6.1. The Mondoñedo thrust sheet

In the allochthonous unit, the recumbent folds reflect
associated to a regional metamorphism of intermediate pressure with kyanite–sillimanite. Little control is available for the prograde paths, but the pressure peaks vary from 6 to 11–12 kbar (Fig. 9a), indicating that the pile of recumbent folds attained a depth of 38–45 km, and the present structural thickness of 10–12 km was initially of 17–20 km (5–6 kbar). After the first Barrovian-type metamorphic event, nappe emplace-

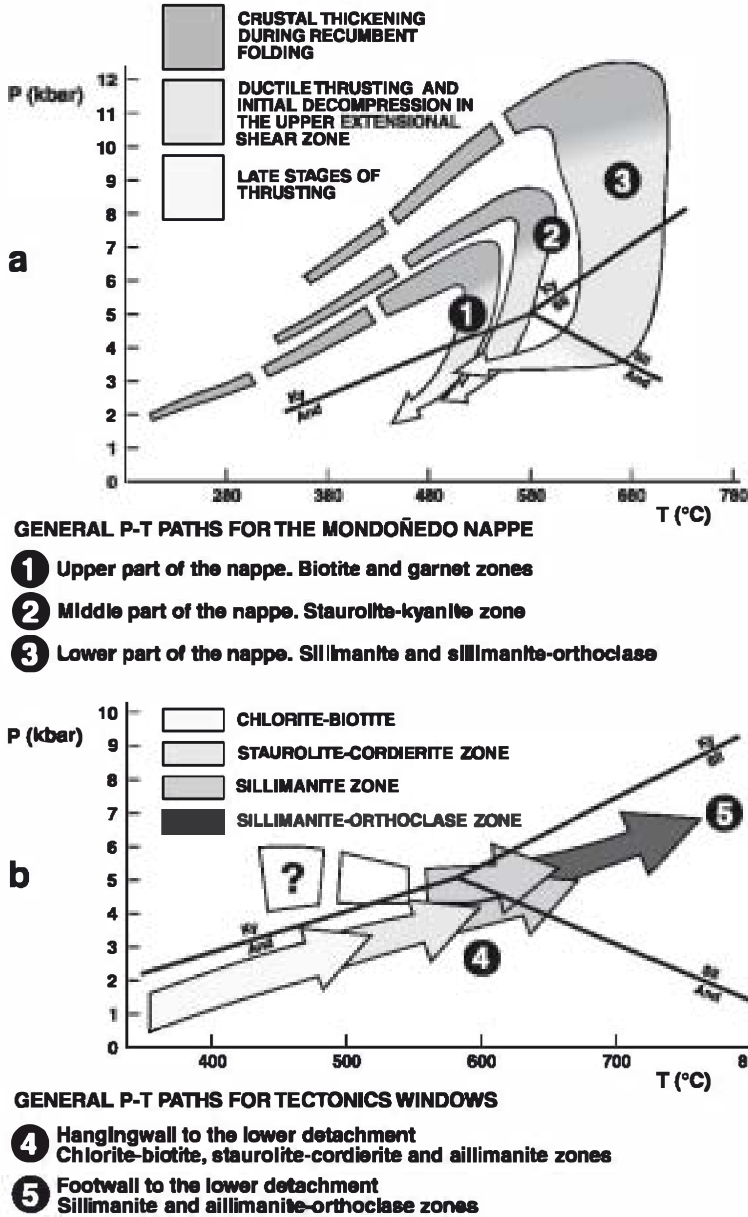

Fig. 9. (a) $P-T$ paths for the upper, middle and lower parts of the Mondoñedo thrust sheet. (b) $P-T$ paths for the footwall unit. The paths are representative of different parts of the nappe and its relative autochthon, which are indicated with numbers in the cross sections of Fig. 3.

ment began along the basal ductile shear zone. Sillimanite grew in the deeper parts of the nappe, whereas andalusite porphyroblasts developed above. All the $P-T$ paths show a decompression, which was greater in the deepest parts (Fig. 9a, paths 2 and 3). The decompressive paths were, at an initial stage, close to isothermal for the basal parts (Fig. 7, paths E to G, path 9a, path 3), suggesting a quick exhumation, a feature commonly associated with tectonic denudation (Thompson and England, 1984). Conversely, the late stage shows cooling accompanied by slight decompression. This corresponds to a thin thrust sheet (3–4 kbar at its base, see path 3 in Fig. 9a), being placed at relatively shallow levels and undergoing only a slight denudation.

It is clear that the thrust sheet underwent
erable thinning during its emplacement, and apparently, the upper extensional shear zone was partially responsible for it. This structure deforms the recumbent folds and the Barrovian zones (Fig. 3). Both the narrowing of the metamorphic zones and the down-dip motion (top to the W) deduced from kinematic criteria are proofs of the subtractive character of the shear zone (Fig. 3). The zone is equivalent to a broad extensional detachment which, in terms of $P-T$ evolution, is reflected
followed by the deep parts of the Mondoñedo

Porphyroblasts of kyanite, staurolite and andalusite are syn-kinematic with a second cleavage in the upper extensional shear zone, implying that the $P-T$ conditions were still high during its activity. This suggests that motion began during the early stages of nappe emplacement. $P-T$ conditions in the extensional shear zone were (at that time) greater than at the base of the thrust sheet in late stages of emplacement. So, the upper shear zone should have finished
before the culmination of the thrusting process. An additional criterion for the relative precocity of the upper extensional shear zone is provided by the Sarria massif. This is a two-mica syn-kinematic granite whose upper part, undeformed, pierced the upper shear zone when its motion had ceased. However, the thrust sheet was moving when the massif intruded on it, because the granite was strongly deformed by the basal shear zone (Figs. 2 and 3, section B-B′).

Though the upper extensional shear zone and the Viveiro fault coincide spatially (Fig. 2), the metamorphic evolution of the Mondoñedo
strates that they were not simultaneous. The upper shear zone developed during early stages of nappe motion, when the thrust sheet was thick and glided above the ductile basal shear zone, whereas the Viveiro fault cuts the thinned thrust sheet, the brittle thrust fault, and the footwall unit, indicating that it formed when nappe motion had ceased. Actually, the Viveiro fault developed under
prints the earlier shear zone, which was dragged downward to the W by the more inclined fault (Fig. 3). Other possibility is that the fault used the pre-existing weak shear zone to nucleate and develop.

The vertical jump between both sides of the Viveiro fault has been calculated in 4 kbar, roughly equivalent to 14 km, by Reche et al. (1998a,b), based on thermobarometry. This throw is viewed as the result of the subtractive motion of the upper extensional shear zone plus that of the Viveiro fault. A dip-slip of 5–6 km is reasonable for the Viveiro fault alone in the N (Fig. 3, section A-A′), which leaves nearly 10 km of throw for the upper extensional shear zone. This figure
structural thickness lost by the thrust sheet, as deduced at the beginning of this chapter form the pressure peaks of their upper and lower parts.

6.2. The Xistral tectonic window

The metamorphic evolution of the footwall unit is better described by its low-pressure conditions and by a strong thermal gradient (Fig. 9b). It is similar to the low-P intermediate type of Miyashiro (1961), characterized by parageneses with andalusite–staurolite–cordierite. An episode of vigorous heating affected most of the Xistral tectonic window, but is specially important above the lower extensional detachment (Fig. 3, section A-A′). The heating produced an abnormal to exaggerated grain growth in the quartzites that are the dominant lithology above the detachment.

High-grade paragneisses of the sillimanite and the sillimanite–orthoclase zones occur below the lower detachment (Figs. 3 and 5). A high-T and low-P penetrative foliation, roughly parallel to the detachment, characterizes its footwall. Numerous granitic injections evidence partial melting, and the metamorphic associations, characterized by the absence of kyanite and the scarcity of garnet, indicate that high temperatures were associated with relatively low pressures (Fig. 9b). No low-T relics are found in the footwall to the detachment, whereas they are common above. In the quartzites of the hangingwall, parallel microinclusions, mostly of white mica, are preserved in largely grown quartz grains, pointing to a first mylonitic stage developed under low-grade conditions. The metapelites in the hangingwall to the detachment are low- to medium-grade schists, which were lately heated up to the sillimanite zone, but never transformed into paragneisses.

Clearly, the detachment brought into contact a low-grade upper unit, above, and a higher-grade unit below. This demonstrates a subtractive jump (Wheeler and

Butler, 1994), coupled with heating of its upper unit with heat transmitted from the lower unit, both features typical of extensional detachments. The heat transmitted from the footwall induced syn- to post-kinematic grain growth in the mylonites of the contractional shear zones above the detachment, including the mylonites of the Mondoñedo basal shear zone preserved in the tectonic window (Martínez Catalán et al., in press). Consequently, the lower extensional detachment was partly synchronous and partly post-dated ductile thrusting. But, as long as the Mondoñedo thrust crosscuts the annealed mylonites, thrust motion must have continued after the detachment activity had ceased.

The $P-T$ paths of the upper and the lower units run almost parallel (Fig. 9b). Both are nearly isobaric, prograde trajectories of the type commonly encountered in the hangingwall to large extensional detachments (Escuder Viruete et al., 1994, 1997). Considering the $P-T$ path of the lower unit (Fig. 9b, path 5), it is possible that this footwall represents in fact the upper part of a deeper zone of crustal thinning, formed by partially melted middle and lower crust and mantle.

Actually, syn-kinematic granites, granodiorites and locally tonalites associated with ultramafic rocks, intruded the northern part of the Xistral tectonic window (Viveiro massif, see Fig. 5). These rocks were deformed, acquiring a low-dipping foliation parallel to the regional foliation of their country rocks. Galán (1987) and Galán et al. (1996) found that the ultramafics
and granodiorites include mantelic components. In addition, the possible existence of mafic
mafic rocks under the Xistral tectonic window is suggested by magnetic and seismic data (Aller et al., 1994; Córdoba et al., 1987; Ayarza et al., 1998).

7. Metamorphism and the structural evolution

The relationships of metamorphism with the structural evolution are sketched in Fig. 10. The successive structural stages are deduced from overprinting criteria among macrostructures, fabrics and mineral assemblages. The tectonic event that formed the large recumbent folds provoked the most important crustal thickening undergone by the Mondoñedo
induced the medium-P metamorphism. The relative delay between the increase of pressure, which is instantaneous, and the increase of temperature, equilibrated the Barrovian upper part of the metamorphic zonation once folding had ceased, as shown by the isograds cutting across the recumbent folds (Fig. 10a and path I, based on Figs. 7 and 9a). The present footwall unit occupied at that time a relatively external and shallow position.

As a response to continued shortening, the contractional ductile shear zones developed. The more internal parts of the nappe were 38–45 km deep when the thrust sheet initiated its exhumation along the basal shear zone (Fig. 10a). Though thrust motion evidently induced continued thickening, crustal thickness was equilibrated by the development of extensional shear zones. The upper one developed first, thinning the rear part of the MLS anticline and the Barrovian metamorphic zones, and balancing crustal thickness in the allochthonous unit (Fig. 10b). Its activity left a clear imprint in the isothermal decompression experienced by the lower parts of the nappe (Fig. 10, path I).

The lower extensional detachment played the same role below the thrust sheet, bringing deep and hot mesocrustal rocks into contact with low-grade supracrustal metasediments. The extensional activity in the lower shear zone had a crucial impact on the metamorphic evolution of the footwall unit. The contractional shear zones, that had previously developed low-grade mylonites, were heated while being still active and continued until annealing once motion had ceased (Fig. 10c and d, and path III, based on Figs. 8 and 9b). Heating was very strong below and above the lower detachment, and syn-kinematic ultramafic
igneous rocks intruded in its footwall. The heat source was probably a layer of partially molten, crustal and subordinate mantelic rocks, trying to open its way upward (Fig. 10, path IV). Its buoyancy may have triggered the development of the lower extensional shear zone.

Once thinned, the Mondoñedo
relatively cold thrust sheet, less than 20 km thick, above a discrete fault. The footwall unit registered a moderate heating and pressurization until the $P-T$ conditions at the top of the footwall were equal to those at the bottom of the hangingwall (Fig. 10, compare paths II and I). Finally, once thrust motion had ceased, the Viveiro fault developed (Fig. 10e), crosscutting the Mondoñedo

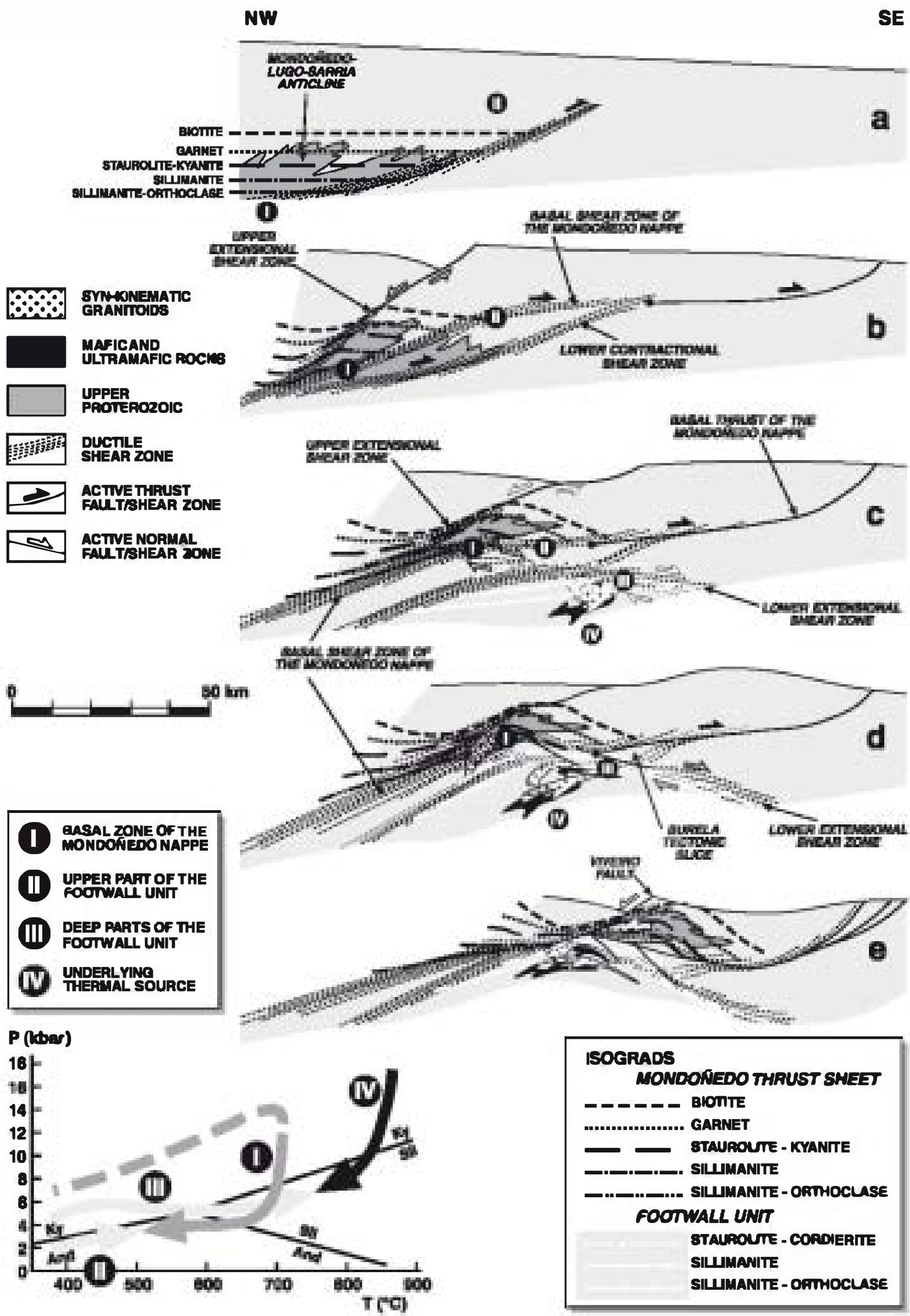

Fig. 10. Structural and metamorphic evolution of the Mondoñedo nappe. See text for explanation.

Available isotopic geochronology permits to constrain the age and duration of the dynamic and metamorphic processes described. These are $^{40}Ar/^{39}Ar$ data obtained from the regional tectonic foliations in the study and surrounding areas (Dallmeyer et al., 1997), and U–Pb data in syn- to post-kinematic granitoids, mafic
Penedo Gordo Massifs (Fernández-Suárez et al., 2000). According to these data, recumbent folding took place roughly between 360 and 340 Ma age, whereas ductile thrusting occurred between 340 and 310 Ma, contemporaneous with the motion on the extensional shear zones. Late stages of thrusting and the Viveiro fault are approximately constrained between 310 and 295 Ma. These aspect are described in detail in Martínez Catalán et al. (in press).

8. Discussion

The metamorphic history of orogenic domains depends of the dynamic evolution of each particular region. However, an extended generalist interpretation tends to ascribe every type of metamorphism to a general kind of dynamic evolution. So, early collisional events are commonly linked to high- and, specially, medium-P metamorphic histories. The late stages of orogenic activity, often marked by crustal extension and abundant magmatism, are related to wide recrystallization in the low-P domain. In that way, both the high-P and medium-P metamorphic types would be of essentially accretionary origin, while the low-P thermal histories would be linked to the gravitational collapse of the tectonic piles built previously. A consequence of this model is that most of the low-P metamorphic belts found in orogenic domains show a complex tectonothermal evolution, where the low-P history tends to overprint and obliterate the previous stages of recrystallization developed under higher baric regimes.

Our research in the Mondoñedo
the generalist models should be used with precaution, and that an accurate interpretation of the metamorphic evolution is only possible after a detailed analysis of the particular dynamic history. A complex structural evolution may explain the development of low-P metamorphism in a context still characterized by convergence and shortening. Moreover, its evolution may show different patterns even in the same region. For instance, it is possible to detect in a stacked sequence the alternation of units where the low-P conditions prevailed during the whole history, with others having undergone a previous medium-P evolution.

The tectonothermal evolution of the Mondoñedo nappe illustrates the complex dynamic evolution of orogenic wedges, where shortening and continued accretion at the base of the pile may be contemporaneous with renewed shortening in the inner parts and with the gravitational collapse of upper and inner levels. This process, masterly anticipated by Platt (1986), has been confirmed
phic and structural analysis of the hinterland of an orogenic region.

Acknowledgements

This work forms part of IGCP Project No. 453, Uniformitarism revisited: a comparison between modern and ancient orogens. Toby Rivers and Cecilio Quesada are thanked for constructive and insightful reviews.

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